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Applicability of thin or thick skinned structural models in a region of multiple inversion episodes; southern South Africa

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Abstract

The deformation of the Cape Fold Belt has been attributed to repeated structural reactivation of a mega-detachment from the late Proterozoic to the Mesozoic (650–65 Ma). Through the integration of onshore cross-sections with observations from the offshore Mesozoic extensional system this study evaluates the applicability of the mega-detachment model.

Regional scale cross-sections through the Permian-Triassic Cape Fold Belt reveal that it comprises two main structural domains: a northern domain dominated by northward verging and asymmetric folds; and a southern domain comprising a series of approximately 8 km wavelength box folds. The genesis of these box folds is attributed to motion on underlying high angle (>45°) reverse faults. This variation between north and south in the fold belt is reflected by a similar variation in extensional geometry of the Mesozoic normal faults, as revealed by subsurface data. The normal faults demonstrate a progressive increase in dip from 24° in the north to 60° in the south.

Features commonly attributed to thin- and thick- skinned tectonic models are observed in both domains, therefore it is not appropriate to describe the observed deformation as one of the two end members. In addition, the structures are inferred to have undergone at least two stages of inversion, irrespective of dip. This is not predicted by either end-member model.

Keywords: Thin-skinned tectonics; Thick-skinned tectonics; Fault reactivation; Structural inversion South Africa

1. Introduction

Two mechanisms are commonly invoked to explain compressional deformation of the continental lithosphere (Fig. 1). Thin-skinned tectonics involves the coalescing of surface thrust faults onto a controlling detachment surface that becomes shallower at depth. Thick-skinned tectonics involve basement, and deformation is principally controlled by the presence and geometry of crustal-scale higher angle structures (e.g. Coward, 1983). Uncertainty in the application of a particular mechanism in many orogenic settings is compounded by the presence of pre-existing structures that may influence the style of compressional deformation through structural inversion. Discriminating between thin and thick-skinned mechanisms, coupled with determining the influence of inversion, can be problematic given available data and the high amount of strain observed in many of these settings. This has led to a number of studies that either re-interpret, or propose contrary models for, the development of orogenic systems (e.g. Coryell and Spang, 1988; Scisciani et al., 2002; Calabrò et al., 2003; Butler et al., 2004).

The Permian-Triassic Cape Fold Belt and the superimposed Mesozoic extensional system of southern South Africa provides a setting in which there is a well-documented (Dingle et al., 1983; de Wit and Ransome, 1992; Hälbich, 1993) long-lived crustal heterogeneity that has undergone a number of structural inversion episodes both in a positive sense (compression utilising extensional structures), and a negative sense

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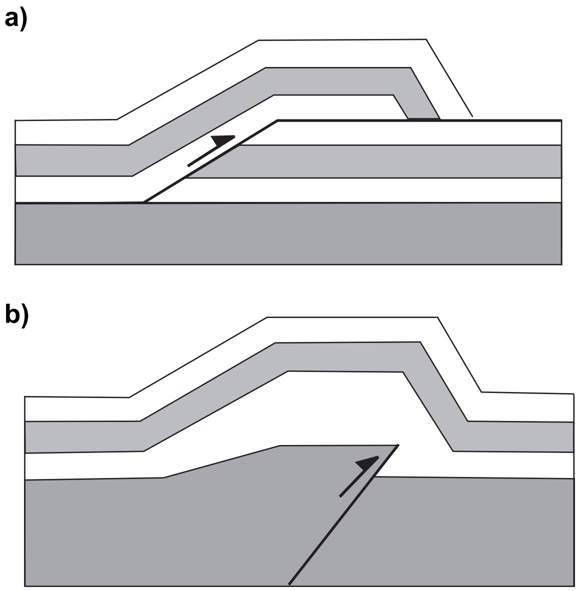

Fig. 1. Cartoons of thin and thick skinned tectonic deformation.

(extension utilising compressional structures; Williams and Powell, 1989). In comparison to many other settings, the Cape Fold Belt structures are reactivated without significant overprinting by subsequent deformation events. Furthermore, the integration of onshore fold belt exposure and offshore sub-surface data provide a significantly greater constraint on fold belt geometry than from surface data alone. Through the application of thin and thick-skinned models, the study area provides a suitable setting to test the applicability of such mechanisms in a relatively undocumented area, and also to evaluate the influence such mechanism have on the structural inversion of a fold belt.

The specific purposes of this paper are three fold: 1) to determine the geometry of the Cape Fold Belt and consider the underlying control on the deformation; 2) to investigate the superimposed extension and use it to constrain the underlying control on deformation; and 3) to consider the Cape Fold Belt within a context of thin- versus thick-skinned tectonics and investigate the influence of structural inheritance.

2. Geological background

It is well documented that the tectonic evolution of southern South Africa comprises a series of extensional and compressional deformation episodes over the last 650 Myrs that are superimposed upon the same crustal heterogeneity (Figs. 2a and 3; Tankard et al., 1982; Dingle et al., 1983; de Wit and Ransome, 1992; Hälbich, 1993; Thomas et al., 1993).

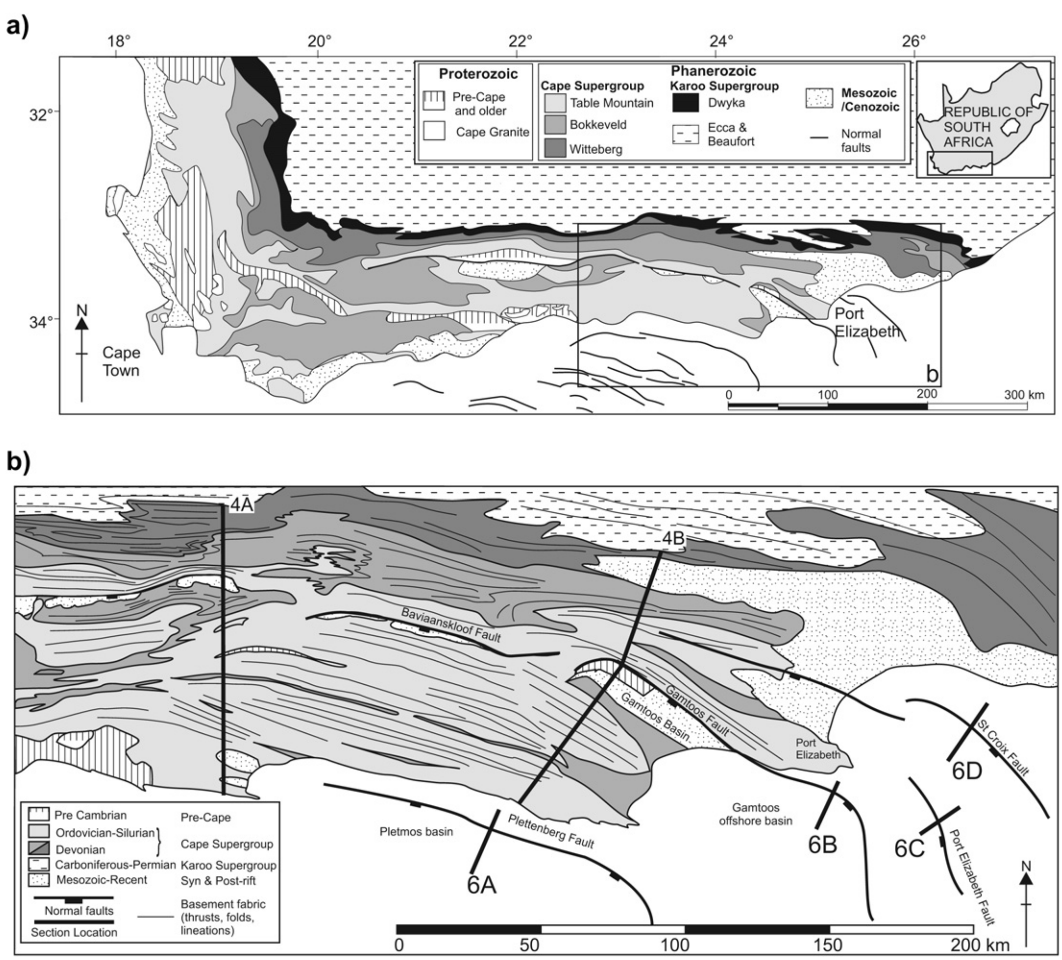

Fig. 2. Geological map of southern South Africa (after Dingle et al., 1983). The location of onshore cross sections (Fig. 4) and offshore seismic sections (Fig. 5) are shown.

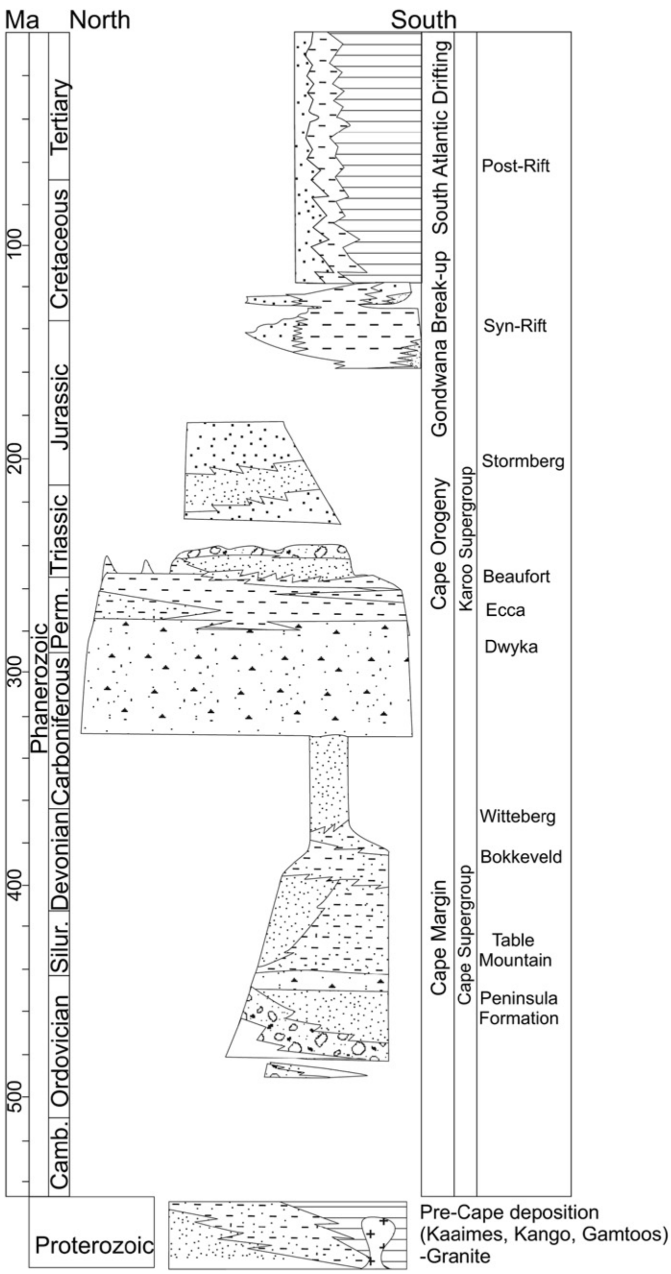

Fig. 3. Chronostratigraphy of southern South Africa with the principal stratigraphic units used in cross-section constructions (after Dingle et al., 1983; Veevers et al., 1994; McMillan et al., 1997; Catuneanu et al., 1998, 2005; Turner, 1999; Booth and Shone, 2002).

One of the principal controls on the development of the region is the Namaqua-Natal Belt that was formed when the passive margin at the southern edge of the Kapvaal craton underwent compression during the Namaqua-Natal Orogeny (950–900 Ma; Thomas et al., 1993). After the cessation of compression, a series of east-west trending extensional basins evolved from 900–600 Ma, within which the Pre-Cape Group sediments were deposited (Fig. 3); these basins were subsequently inverted during the Pan African Orogeny (600–450 Ma, Tankard et al., 1982; Gresse, 1983; Krynauw, 1983; Shone et al., 1990). Subsequent to the Pan African Orogeny an intra-continental clastic margin was established onto which the Ordovician to Early Carboniferous Cape Supergroup was deposited. The development of the Gondwanian orogeny in the Permian resulted in the termination of Cape Supergroup deposition. The Gondwanian deformation formed a fold belt that is traceable from the Sierra de la Ventana in Argentina to the Pensacola Mountains of the Trans-Antarctic Mountains (Du Toit, 1937; Dalziel et al., 2000). In South Africa, compression was manifested through deformation of the Cape Supergroup margin deposits into the Cape Fold Belt and formation of the Karoo foreland basin to the north (Fig. 2; Hälbich, 1983; Hälbich, 1993; Veevers et al., 1994).

During the break-up of Gondwana, and subsequent rifting of the South Atlantic, the Cape Fold Belt underwent negative structural inversion, resulting in the superimposition of a Mesozoic extensional system onto the fold belt (De Wit and Ransome, 1992; Hälbich, 1993). Rifting is considered to have initiated in the Middle Jurassic, with deposition of terrestrial and shallow marine sediments (Dingle et al., 1983; McMillan et al., 1997). Shallow or non-marine sedimentation continued for the entire rift episode onshore while the rate of extension rapidly increased in the offshore portions resulting in an abrupt transition to a deepwater setting that continued for much of the rift episode (McLachlan and McMillan, 1976; Shone, 1978; Dingle et al., 1983; Paton and Underhill, 2004). The rift-drift transition is considered to be Valanginian in age and is overlain by post-rift shallow marine sediments (McMillan et al., 1997).

3. Compressional regime

3.1. Cross-section construction

Cross-sections were constructed from original fieldwork Dingle et al., 1983; Veevers et al., 1994; McMillan et al., 1997; Catuneanu et al., 1998, 2005; Turner, 1999; Booth and Shone, 2002). data, ETM+ satellite image interpretation, and published geological maps (Republic of South Africa Survey maps 3320, 3420, 3322, 3324), and extend from the Karoo Foreland basin in the north across the Cape Fold Belt to the southern coast (Figs. 2 and 4). Sections were oriented perpendicular to the structural trend of the fold belt (west-east in the Central Cape and NW-SE in the Eastern Cape) and were constructed using standard techniques of line length and area preservation (Dahlstrom, 1969; Elliott and Johnson, 1980; Boyer and Elliott, 1982; Elliott, 1983). The geometry, dip, and continuity of sequences at the surface were used to derive fold geometries and fault locations. Folding was considered to be either generated through buckling above an underlying décollement horizon, or associated with a forcing controlling fault. Fold trains with regular spacing, equal amplitudes, low aspect ratios and wavelength as a function of unit thickness, were assumed to be associated with buckling (e.g. Morley, 1994; Mitra, 2002); while asymmetric, high aspect ratio folds were considered to be fault forced folds. In the latter, the location of the underlying controlling fault, if it was not present at the surface, was inferred from fold-limb dip and fold axis location (Boyer and Elliott, 1982; Mitra, 1990, 2002). Sequence

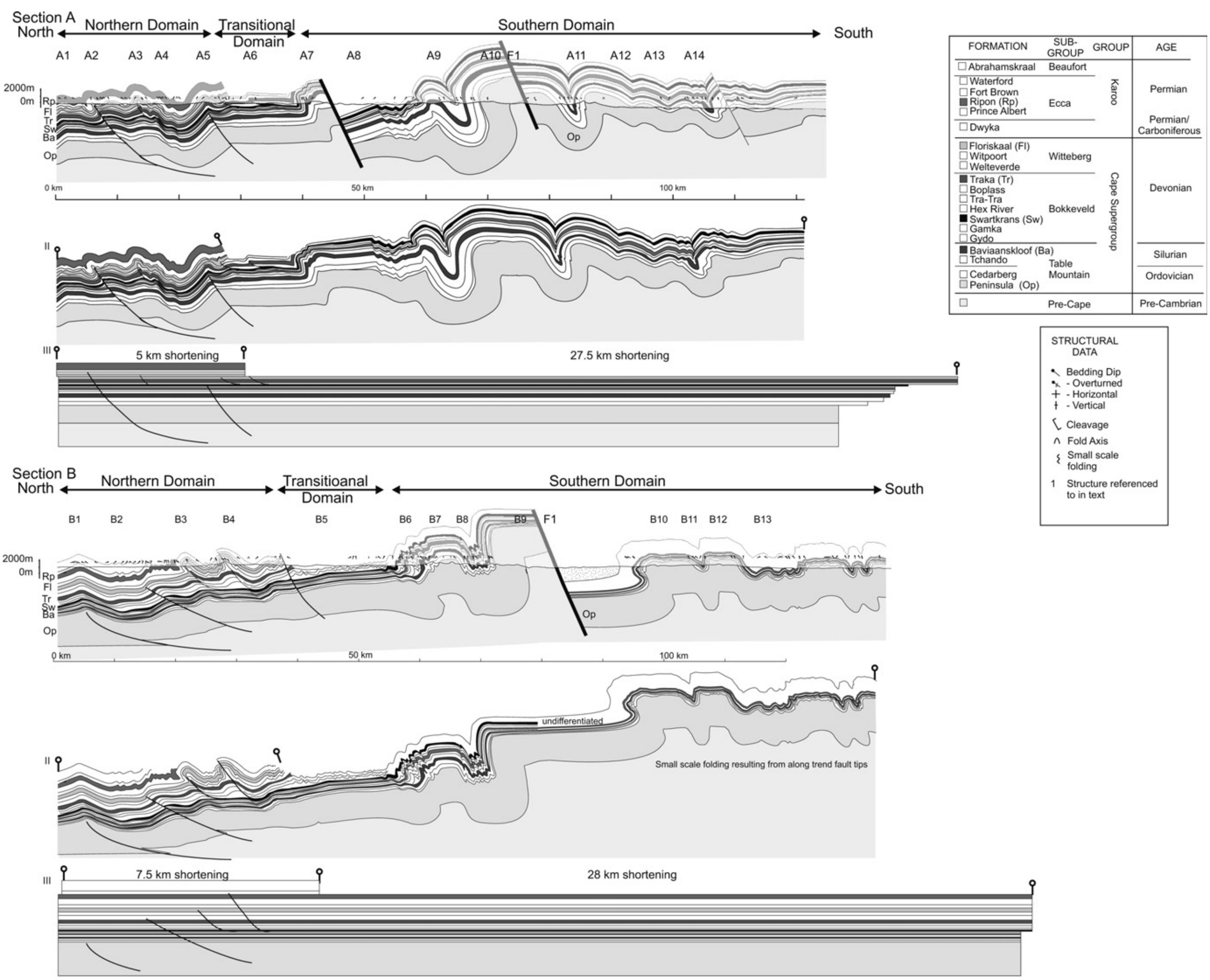

Fig. 4. Semi-balanced cross sections across the Cape Fold Belt (Fig. 2 for location); see text on discussion of why sections have not been fully balanced. Numbers in cross section represent structures referred to in the text. For each section: I) present day geometry with topography and structural data plotted; II) restoration of Cape Fold Belt through the removal of extensional faults; and III) restored sections to pre-Cape Fold Belt deformation. Sections have been constructed with as little inference of underlying fault geometries as possible, and with thickness continuity of the Cape Supergroup succession (see text for discussion).

thicknesses were derived from locations where structural data were abundant and sequence geometries were well constrained. These thickness were found to be consistent both along and between sections and to agree with published data (Toerien, 1979; Toerien and Hill, 1989). This consistency in sequence thickness was used as a further constraint on extrapolated geometries. As a number of previous studies have documented the approximate plane strain nature of the deformation (Hälbich, 1993; De Wit and Ransome, 1992; Söhnge and Hälbich, 1983) errors and uncertainties associated with-out-of plane transport were considered to be negligible. The extrapolation of surface data presented here does not necessarily present a unique solution; however, it represents a consistent interpretation of the data with as limited an extrapolation as is permissible. Section balancing was performed using Midland Valley 2D Move software (Version 3.1) in which a number of horizons were restored to a pre-deformation state. The initial restoration stage involved the removal of extensional displacement on the normal faults through the rotation of hangingwall geometries to a pre-extensional configuration (Fig. 4). It was assumed that faults were pure-dip slip and that hangingwall and footwall blocks acted as rigid blocks with no internal deformation. The resultant pre-extensional geometry was then unfolded by restoring four target horizons (top Ripon Formation, Middle Ecca, Permian; top Witteberg Group, Middle Devonian; top Bokkeveld Group, Devonian/Silurian; and top Peninsula Formation) to the horizontal using conservation of bed length.

3.2. Stratigraphy

The stratigraphy of the area comprises over 3000 m of meta-sediments of the Pre-Cape Group, although these only crop out as occasional inliers that lie unconformably below approximately 8 km of Ordovician to Early Carboniferous Cape Supergroup clastic margin sediments and up to 9 km of Permian and Triassic Karoo foreland basin deposits (Fig. 3; Toerien, 1979; Tankard et al., 1982; Veevers et al., 1994). With the exception of the Pre-Cape phyllitic hornfels and schist, the succession has only undergone very low-temperature and low-pressure metamorphism, with a gradual transition from unmetamorphosed Karoo sediments through an upper anchimetamorphic zone (T ca. 150°, $P = 23$ kb) to lowermost greenschist facies (T ca. 350 °C) on the south coast (Hälbich and Cornell, 1983; Duane and Brown, 1992).

The lowest sequence of the Cape Supergroup is the Early Ordovician to earliest Devonian Table Mountain Group, which is predominantly cross-bedded, super-mature, medium to coarse-grained, quartzitic sandstones, conformably overlain by argillaceous fine grained sandstones and more massive shales and siltstones of the Early to Late Devonian Bokkeveld Group (Toerien, 1979; Bell, 1980; Toerien and Hill, 1989; Booth and Shone, 1999). The uppermost Cape Supergroup comprises shales and subordinate sandstones of the Late Devonian to Early Carboniferous Witteberg Group.

The sediments of the Karoo Basin are sub-divided into the Dwyka, Ecca, Beaufort and Stormberg groups and have a total thickness of up to 9 km (Toerien, 1979; Toerien and Hill, 1989; Catuneanu et al., 1998, 2005; Turner, 1999). The Dwyka tillite is disconformable on top of the Witteberg Group; the first episode of the Cape Orogenic compression occurred during the lithification of this unit (Hälbich and Swart, 1983; Cole, 1992; Veevers et al., 1994). The Ecca Group unconformably overlies the glacial deposits and comprises sandstones, carbonaceous shales and limestones while the Beaufort Group of the Early Triassic is dominated by mudstones and sandstones of a meandering river system that is the consequence of a large-scale regression (Dingle et al., 1983; Cole, 1992; Turner, 1999). The Stormberg Group overlies the Beaufort Group and comprises the proximal, fluvially dominated Molteno Formation and the distal Elliot Formation (Tankard et al., 1982; Turner, 1986; Cole, 1992; Turner, 1999). The Karoo flood basalts that cap the foreland basin sedimentary succession are not present in this study area (Tankard et al., 1982).

3.3. Structural domains

The regional cross-sections reveal a consistency in fold belt geometry along the trend of the structural fabric (Fig. 4). There is, however, a clear difference in deformation character between the northern and southern structural domains. The transitional domain, between the two, is co-incident with a change in topography from the moderately low relief Karoo basin to the higher relief of the principal fold belt area.

3.3.1. Northern structural domain

The northern structural domain is dominated by the low-relief Karoo foreland basin composed of deformed Carboniferous and Permian age tillite and turbiditic strata. The magnitude and style of deformation across both sections represent a spectrum of deformation from microscopic to km-scale wavelength folds and symmetric to highly asymmetric fold limbs.

In a number of localities reverse faults crop out at the surface and are coincident with the surface axial traces of highly asymmetric, tight to isoclinal (inter limbs angles of <40°) anticlines with moderate (~50°) south dipping southern limbs and steep (>70°) to overturned northern limbs (for example folds A1, A3 and B4). The development of these folds is considered to be intimately linked to the faults observed at their fold axes. In the modeled cross sections, in order to prevent space and area preservation problems during cross-section construction, the faults associated with these folds are inferred to have greater displacement at depth. None of the modeled faults have significant offsets on the surface, and the cross sections predict that they do not have displacement greater than approximately 3 km at depth.

Other surface folds (e.g. A5, B3) have very similar fold limb geometries, amplitudes and wavelengths, and although faults are not evident at the surface axial trace of the folds, they are considered to be fault propagation folds in which the underlying fault displacement reduces up dip with shortening being accommodated by folding in front of the fault tip (Mitra, 1990). In the fold hinges of some of these observed folds, most noticeably folds A3 and B3, the outcrop width of the lower Permian sequence is wider than that predicted from the unit thickness, given the local structural constraints at the fault hinge (for example fold A3 is 420 m wide instead of 250 m), implying thickening at some fold hinges.

In contrast to the asymmetric, fault forced folds, fold B1 is an open fold (inter limb angle 220°) with symmetric fold limbs and no significant faulting within the fold hinge. The extrapolation of the concentric geometry to the sub-surface results in significant space problems at depth in the cross section models and can only be accounted for by a buried fault, or thickening of a lower sequence within the fold hinge above a detachment layer. From the current data, in particular the symmetrical nature, it is inferred that the fold is not fault generated; and by assuming that the amplitude of the structure is approximately the same as the layer thickness of the deforming package, the detachment layer is considered to be in the upper Cape Supergroup.

In addition to the large folds, there is a population of smaller scale folds (wavelengths on order of ~100 m) that are either in regions of little deformation between longer wavelength folds (e.g. A4, B2), or are superimposed upon the limbs of the larger folds (e.g. A4, B3). Given that these folds are commonly symmetric they are considered to be folds associated with localised detachments. The low amplitude character of the folds implies that the controlling detachments are shallow, and as these folds are evident across the structural domain, regardless of the exposed sequence, it is inferred that there are detachment horizons throughout the Karoo Supergroup. This is consistent with the inter-bedded sandstone and shale lithologies of much of the turbiditic Karoo sequences.

In both sections the southern extent of this domain corresponds to the most southerly outcrop of the Dwyka tillite. As this formation has been demonstrated to have a syn-depositional relationship to the first phase of Cape Fold Belt deformation, previous workers have concluded that this corresponds to the southernmost limit of Dwyka, and lower Karoo Supergoup sediment deposition (Hälbich and Swart, 1983; Cole, 1992). In the sections in this study, the Karoo Supergroup has therefore not been extended farther south than this location. In this structural domain, neither section requires restoration to account for Mesozoic extension, therefore section balancing is undertaken only to remove the compressional phase. When fold and fault geometries are extrapolated to depth, and accounting for an increase in fault displacement with depth, the upper sequences can be balanced. The lower sequences, in particular the Peninsula Formation, are difficult to balance and therefore an additional fault is inferred that is manifested in the surface of the section by fold A2. Although some of the folds and faults are compatible with a relatively shallow décollement level, i.e. within the upper Cape Supergroup, certain features, such as the syncline to the south of fold A4, require a detachment at least at the base of the Peninsula Formation;, therefore, it is inferred that the common detachment horizon is at this depth. I order to balance the section there is no requirement for duplexing of sequences, or for large scale detachment of allochthonous units. This is consistent with there being no significant out-of-sequence outliers or thrust-bound klippen. The estimated amount of shortening for the northern structural domain is similar in both sections with 5 km in Section A and 7.5 km in Section B. The pin lines from which these values are calculated are along trend of each other.

3.3.2. Transitional structural domain

Immediately to the south of the northern domain, there is a consistent outcrop of upper Devonian strata with a width of 15–20 km that is characterised by abundant short wavelength (<100 m), low amplitude (<50 m) folding (Folds A6, B5; Fig. 4). In this domain, only the lower Witteberg Group crops out. The scale and symmetric geometry of fold limb dips indicate buckle folding and as there are only two formations involved and the amplitudes are small, it is inferred that there is a very shallow décollement horizon. As there is negligible shortening across this domain there is no requirement for deeper level shortening to be included within the sections.

3.3.3. Southern structural domain

The southern structural domain forms the dominant topography of the southern Cape with elevations of up to 1500 m. Stratigraphically, it is formed by the Cape Supergroup with the quartzitic and arenaceous sandstones of the Peninsula Formation, which forms the highest topography, and the more easily eroded Bokkeveld and Witteberg subfeldspathic and sublithic sandstones and mudrocks.

In both sections the northern part of the domain is characterised by exposure of the entire lower Devonian to Ordovician succession with steep (>80°) to overturned dips and no significant structural discontinuity; this forms the northern limb of the Peninsula-cored anticlines (locations A7, B6). Such anticlines dominate the Cape Fold Belt and are separated by structurally well-constrained Silurian and occasionally lower Devonian cored synclines.

The principal limitation of regional studies of the Cape Fold Belt is determining the internal structure of the Peninsula Formation exposure because of localised structural complexities, including syn-depositional soft sedimentary deformation (Hälbich, 1983) and the absence of easily correlatable intra-formation horizons. Despite these complications, the regional-scale deformation features of steeply dipping fold limbs with complex folding and faulting deformation between limbs are in agreement with a number of studies that have undertaken detailed analysis of portions of the fold belt (Söhnge and Hälbich, 1983; De Wit and Ransome, 1992).

The outcrop of the Peninsula Formation consistently comprises three structural portions. The northern portions (Fig. 5a) consist of steep (>70°) to overturned beds that are, at a regional scale, conformable with the steeply dipping Devonian and Silurian sequences. This transition in Section A (A7) is relatively undeformed with a consistent overturned dip (~80°) towards the south, while in Section B (B7) there is more variability of dips ranging from 60° to vertical or overturned. In the latter section, the Devonian sequence is also characterised by complex small scale folding, which corresponds to the disharmonic folding described by Hälbich (1983). This disharmonic folding may be equivalent to the buckle folding described in the Transitional structural domain. The central section of the Peninsula Formation exposure consists of a complex zone of deformation commonly with extensive chevron folding (Fig. 5b). Deformation is accommodated through the folding of ~50 cm thick quartzitic sandstone layers with flexural slip occurring in the interbedded siltstone dominated layers. To the south of the chevron folding there is a zone of little or no deformation with sub-horizontal, or gently south-dipping bedding (Fig. 5c). In a number of the Peninsula Formation anticlines, the central core consists of exposure of the Pre-Cape units. The structure of the Pre-Cape units is commonly very complicated with abundant folds and thrusts with displacements of up to 3 km. Where the deformation has been well documented (e.g. Gresse, 1983; Fig. 5d), the style of deformation, including thrust faulting, resembles that of the triangular deformation zone above basement faults discussed by Mitra and Mount (1998). The sub-horizontal Peninsula Formation, or where present the Pre-Cape units, in both sections are commonly dissected on the southern edge by a south dipping normal fault (e.g. A7, A10, & B9). This observation is consistent across the fold belt with the southern portion of many of the box-folds being down-faulted towards the south by normal faults (Gresse et al., 1992). The extensional faults will be discussed in more detail in the next section.

To the south of the non-deformed zone, or the normal faults, there is commonly a zone of steeply south-dipping (>70°) strata with little deformation (Fig. 5e). This south-dipping limb commonly includes the lower Devonian to

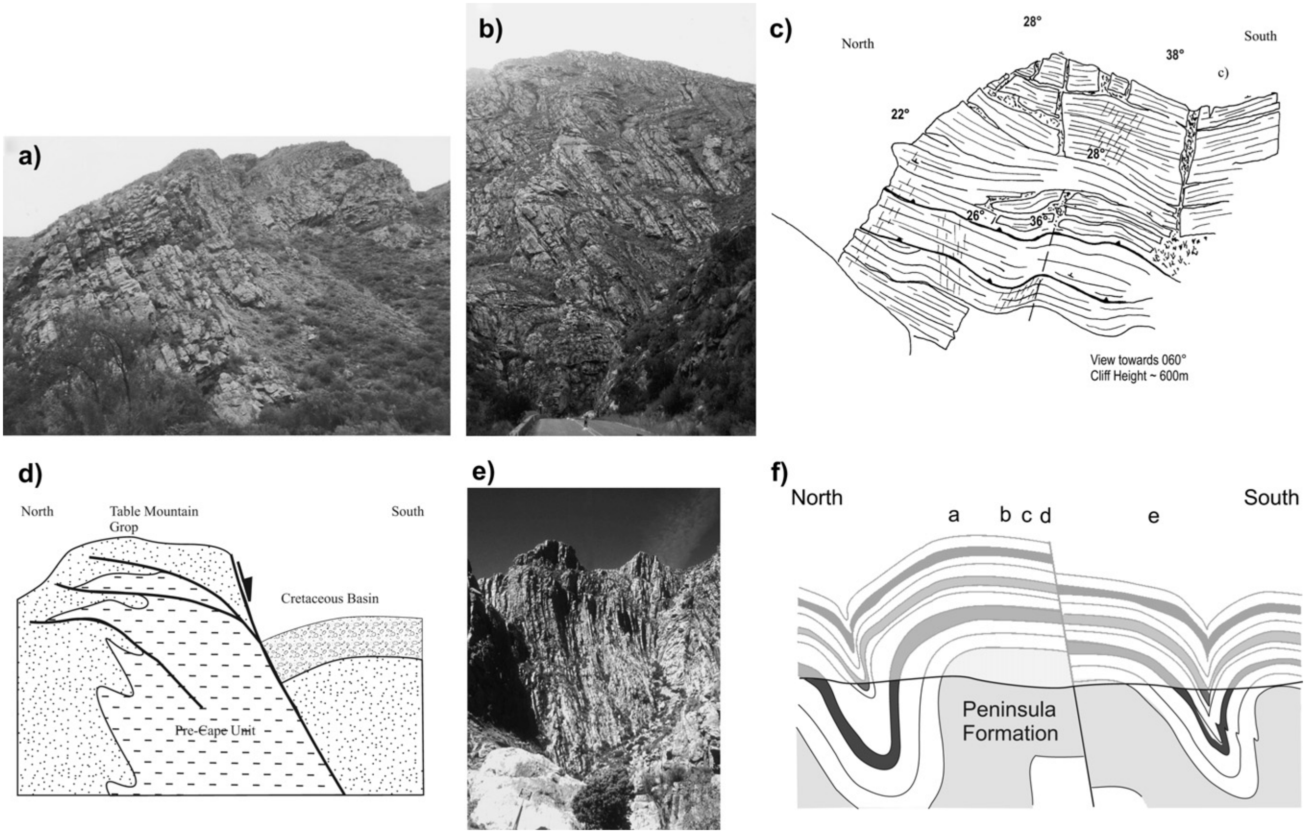

Fig. 5. Photographs and field sketches demonstrating the change in deformation within the Peninsula Formation units from north to south: (a) steeply dipping to overturned bedding planes in the north: (b) chevron folding; (c) central portion dominated by sub-horizontal to gentle south dipping beds; (e) deformation of the Pre-Cape units (after Gresse, 1983), that resemble triangular deformation associated with high angle basement faults (Mitra, 1990); (d) steeply dipping beds. Overall the Peninsula Formation represents box fold structures.

Peninsula Formation sequence and corresponds to the northern limb of the Devonian/Silurian-cored synclines.

On a regional scale, therefore, the Peninsula Formation and Pre-Cape exposures correspond to anticlinal box folds with steeply dipping limbs and sub-horizontal or gentle southerly dipping internal portions. This Peninsula Formation box fold geometry is repeated across both sections with amplitudes of approximately 5000 m although wavelengths vary from 2000 m (B7) up to 8000 m (A10, B9). This study has only considered the geometry of these folds in two dimensions, however, plan view geometries demonstrate that these structures are not continuous along the entire length of the fold belt (Fig. 2b, Dingle et al., 1983) and periclines are common. The occurrence of the shorter wavelength folds are co-incident with such features and are, therefore, considered to be close to the termination of the folds (e.g. A14, B11 & B13). The box folds in the sections have very little structural elevation across them, although folds A10 and B9 are the exception as they have significant structural elevations in addition to greater amplitudes and wavelengths. In the two sections this corresponds to the same structure.

As discussed, the box folds are consistently separated by lower Devonian and Silurian strata that are exposed in northward verging, tight synclines (commonly northern limb has dip of ~75° with an overturned southern limb). The synclines have similar amplitudes to the box folds (~5000 m) although have significantly shorter wavelengths (~4000 m).

Although within the sections presented here, no thrust faults have been observed, a number have been identified occurring within the triangular deformation zone as discussed by Gresse et al. (1983).

The balancing of the sections is problematic because Pre-Cape unit geometry and thickness cannot be constrained. A simple restoration of the fold belt, assuming consistency of sequence thickness, results in an area deficit for the lower sequences, in particular the Peninsula Formation. The amount of thickening associated with chevron folding is difficult to quantify and accounts for part of the deficit (Fig. 5b). In addition, the application of a uniform thickness to the Peninsula Formation is not necessarily appropriate as will be discussed later.

4. Extensional regime

4.1. Methods and data

In addition to the onshore extensional geology, which has been included within the cross-sections, there are a series of offshore extensional basins that can be used to better understand the relationship between extension and compression.

The offshore component utilised 19,000 km of 2D seismic reflection profiles tied to 41 exploration wells and covering 23,000 km^2 made available by the Petroleum Agency South Africa. All seismic data have a vertical axis in milliseconds two-way-travel-time (ms TWT) with maximum recording values of either 5000 or 6000 ms, 60 fold geophone coverage, and 25 m shot point interval. Depth conversion of key sections from TWT to depth in metres was undertaken using published data (McMillan et al., 1997; Paton and Underhill, 2004). Through the application of standard techniques of seismic facies and reflection termination identification, the top basement reflection was picked and used to define the cross-sectional geometry of the basin-bounding faults (Fig. 6). The top basement reflection corresponds to a high amplitude double wavelet that separates the semi-transparent basement reflection character with the concordant, or onlapping, reflections of the syn-rift interval. In addition, where the top basement pick was penetrated by boreholes it conforms lithologically to the quartzitic sandstones of the onshore basement.

4.2. Correlation of extensional and compressional regimes

The onshore extensional component of southern South Africa comprises a series of Mesozoic half graben that are bound to the north by controlling normal faults (Fig. 2; McMillan et al., 1997). A number of studies have demonstrated the correlation in trend between compressional and extensional structures and demonstrated that the extensional faults re-activate the compressional faults in cross-section (De Wit and Ransome, 1992). The principal normal faults have an east-west trend (pole to the π-girdle of 00.2° to 096°) in the central Cape and a NW-SE trend in the eastern Cape (pole to the π-girdle of 07° to 133°).

The cross-sections presented in this study intersect a number of normal faults and the dips of two of these faults (AF1 & BF1) are 60° towards the south. Although Mesozoic sediments show divergence into the faults there is little control on the amount of displacement on the faults except for the Gamtoos Basin in which two boreholes penetrate at least 2326 m of Mesozoic sediments (wells Mk1/70 and Lo1/69). Elsewhere in the southern Cape previous studies have estimated that displacements are approximately 8 km (Dingle et al., 1983). The normal faults in both of the sections occur in a structurally consistent location immediately to the south of the northern limb of the anticlinal box folds. A number of other studies have demonstrated that this is consistent across the fold belt and that all normal faults occur in this position (Gresse et al., 1983; Dingle et al., 1983; Paton, 2006).

4.3. Subsurface extensional geometry

The geometry of the sub-surface extensional system is derived from the available offshore seismic data within the Pletmos, Gamtoos and Algoa basins (Fig. 2). They all have a half-graben geometry with basin-bounding faults in the north and north-east that dip towards the south and south-west. The trace of the faults are parallel both to the immediately adjacent onshore basement structure and to the regional gravity data (Dingle et al., 1983; McMillan et al., 1997; Paton and Underhill, 2004; Paton, 2006). The high resolution of the seismic data coupled with the large impedance contrast between the siltstone-dominated syn-rift and quartzitic basement result in a well defined location for the fault planes (Fig. 6). In some portions of the section the data quality is reduced by diffraction (a consequence of incomplete migration) and occasional refraction of syn-rift reflection ray paths through the fault plane, although it is still possible to locate the fault plane with confidence.

The Plettenberg Fault, which is the controlling fault in the northern Pletmos Basin, has a dominantly east-west strike, although there is a change to a north-south strike in the east of the basin. The fault is imaged to at least 5500 ms, which is equivalent to 12,000 m, has a planar geometry for its entirety with a dip of 65° after depth conversion (Fig. 6A). The fault is immediately offshore of the southern extent of Section B.

The Gamtoos Fault has a very similar geometry to that of the Plettenberg Fault, and although it also has an east-west trend, the north-south trend is more dominant at the eastern margin of the basin. In cross-section, the fault plane has a planar geometry to at least 5000 ms, equivalent to 11,000 m, a depth converted dip of 42.5°, and a displacement of 16,500 m (Fig. 6B). The offshore fault is a direct continuation of the Gamtoos Fault that dissects Section B.

Although the Port Elizabeth Fault (Fig. 6C) has a similar planar cross-sectional geometry to that of both the Plettenberg and Gamtoos Faults, it has a shallower dip (37°) and the true heave can not be determined because of a significant basin-scale unconformity. The top basement pick in the hangingwall occurs at a more shallow depth than the previous two faults at 4000 ms and has a minimum displacement of 8600 m.

The geometry of the St Croix Fault (Fig. 6D) is significantly different from the other faults because, although it has a depth converted displacement of 12,600 m, it has an average dip of 24° and has a listric geometry that flattens out at an approximate depth of 8000 m. Although this section does not correspond directly to a position on either of the sections, it is directly along trend of fold B5.

5. Controls on Cape Fold Belt geometry

5.1. Integrating onshore and offshore observations to determine subsurface geometries

The occurrence of northward verging, commonly tight, asymmetric folds and occasional thrusts in the northern domain is consistent with a series of moderately shallow, south dipping thrust faults that are considered to coalesce at depth onto a common décollement. Given the geometries of the folding, in particular features such as the syncline to the south of fold A4, it is likely that the principal décollement is below the Cape Supergroup and within the Pre-Cape Group; a minor detachment is present within the Upper Devonian. The St Croix

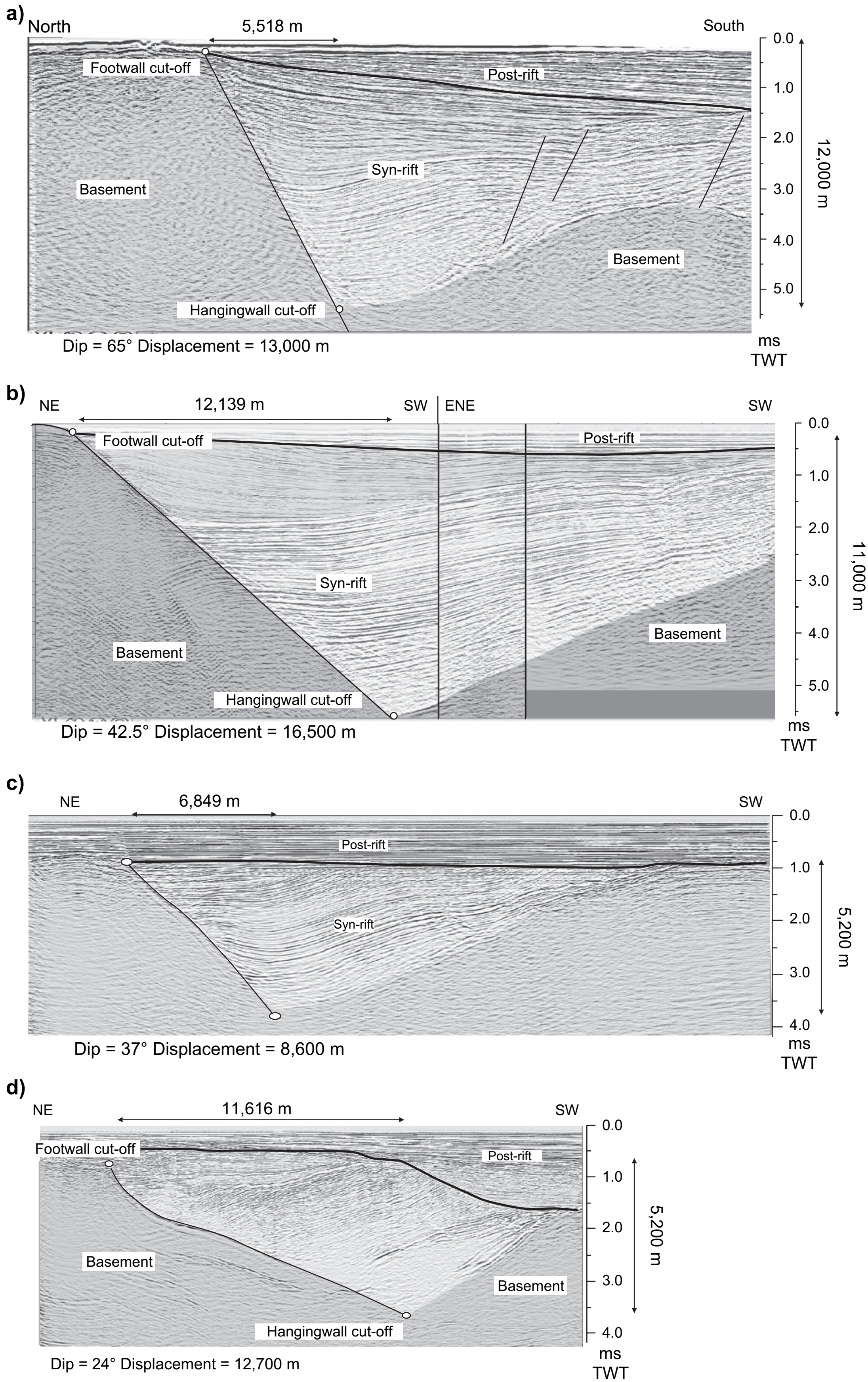

Fig. 6. Seismic sections across the principal extensional faults from north to south: (A) Plettenberg; (B) Gamtoos; (C) Port Elizabeth; and (D) St Croix. Sections have been depth converted and demonstrate a sequential decrease in dip from south to north. See Fig. 2 for locations.

Fault, which is along strike of the northern domain, is a moderately shallow (24°) dipping fault that flattens at a depth of approximately 6000 m. The St Croix Fault, therefore, has a geometry consistent with the compressional faults derived from the cross-sections and is considered to be an extensionally reactivated reverse faults.

The box folds separated by relatively narrow synclines within the southern domain resemble buckle folding above an underlying detachment (e.g. Costa and Vendeville, 2002). As the Pre-Cape units are present in the core of some of the box folds this would suggest that any detachment would have to be within the Pre-Cape sequence, and hence below the Cape Supergroup. Such buckling is problematic for three reasons. Firstly, the thickness of the Cape Supergroup is at least 8 km, and as the décollement would have to be within the Pre-Cape unit it is likely that it is not shallower than 8 km. Such an assertion is also supported by the amplitude of the box fold, assuming that the amplitude is a function of deformational unit thickness (Mitra, 2002). Secondly, most documented examples of box folds forming above detachments are shallow features commonly associated with salt and are, therefore, not consistent with the Cape Fold Belt setting. Thirdly, most models and examples of buckling have no significant structural elevation across the section, however, in the sections presented in this study there is a significant structural elevation across the Kango Fault area.

An alternative process for generating box folds is having the folds as fault generated. Although most examples of fault forced folding results in asymmetric, tight folds, as is observed in the northern structural domains, a number of studies have demonstrated that the dip of the controlling fault has a significant influence on the style of deformation of the overlying sequence. Bonini et al. (2000), using sand box models, demonstrated the intimate link between ramp angle of a thrust sheet and the geometry of the overlying anticline. Shallow-dipping ramps of 15–30° resulted in long-wavelength, low-amplitude anticlines, while high-angle ramps of 45–60° produced lower wavelength and higher amplitude anticlines that closely the resemble the box folds of the Cape Fold Belt. These results are in agreement with a number of other sand box studies, including Buchanan and McClay (1991) and McClay (1995), which predict that if the controlling fault is steeply dipping (45°–60°), a box fold with an antithetic backthrust would be generated. Kinematic and numerical modelling also predict box fold generation only from steeply dipping controlling faults (e.g. Salvini et al., 2001; Savage and Cooke, 2003). These models

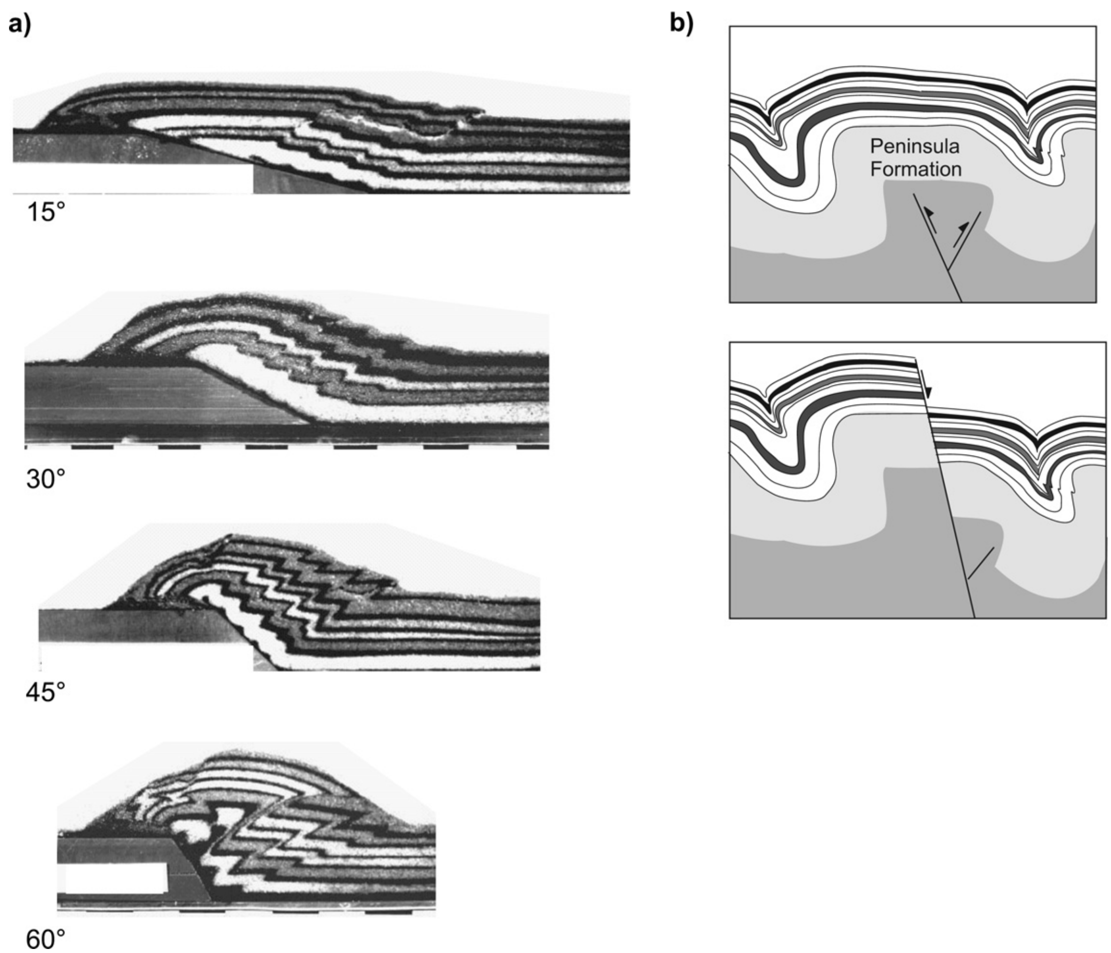

Fig. 7. a) Results of sand box modelling demonstrating the change in anticlinal geometry associated with increase in dip of underlying ramp (after Bonini, 2000). b) Application of sand box models to the box folds observed within the structural domains and the observed location of the Mesozoic extensional faults.

predict that the location of the controlling fault would be associated with the steeper dipping fold limb, hence in this study area they would be directly to the south of the northern box fold limbs (Fig. 7b).

The presence of the onshore Mesozoic extensional faults consistently to the south of the northern box fold limbs, with dips of approximately 60°, is therefore consistent with the fault controlled box fold model if the extensional faults represent structurally inverted high angle reverse faults. Not only do the offshore faults support this premise, their geometry also demonstrates that these high-angle inverted faults are planar structures that continue to at least a depth of 12,000 m with a steep dip (Fig. 8).

5.2. Implications for the crustal-scale control of the fold belt and its inversion

Although the structure of the middle and lower Cape Fold Belt crust remains poorly understood (Harvey et al., 2001), a number of previous studies have proposed the presence of a south-dipping mega-detachment surface (Hälbich, 1993; De Wit and Ransome, 1992; Söhnge and Hälbich, 1983).

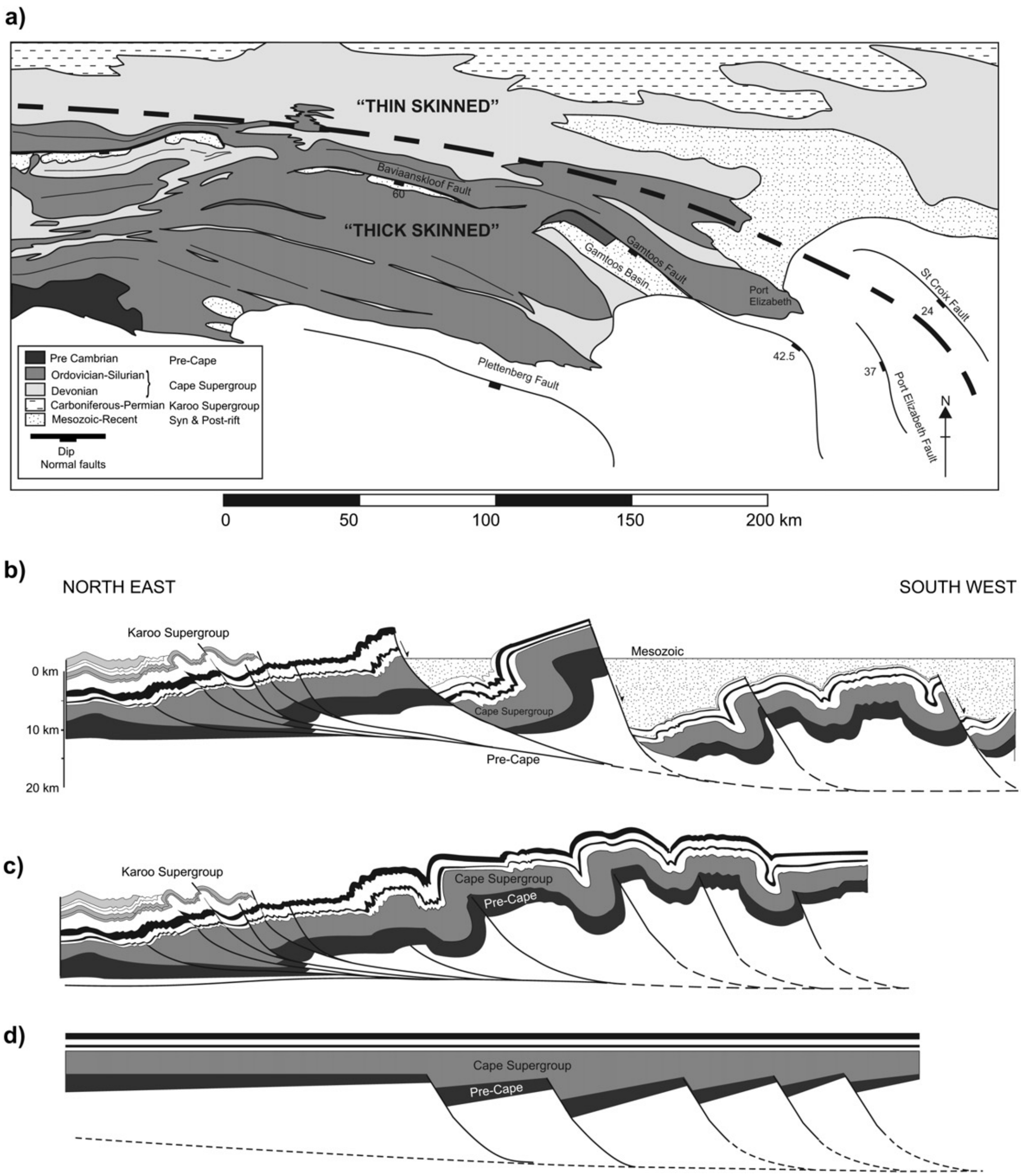

Fig. 8. a) Distribution of the normal faults, and fault dips, with respect to the characterisation of the Cape Fold Belt compressional deformation into ''thin-skinned'' (northern) and ''thick-skinned'' (southern) domains. b–d) Model for the development and fault inversion of the Cape Fold Belt, demonstrating the repeated reactivation of high angle faults in the south and lower angle faults in the north. The depth to the décollement is constrained by the depth converted sections in Fig. 6. It is proposed that the high angle faults that formed the box folds (c) were active as normal faults during the Cape Supergroup passive margin (d).

Many of these studies propose that this feature is reactivated during sequential episodes of deformation. However, none of these studies have fully integrated the offshore observations with those from onshore.

The observations presented in this study of a south dipping detachment in the northern domain are consistent with a mega-detachment. In the models presented in this study the detachment is inferred to have a depth of approximately 8 km at the southern edge of the Karoo Basin; this is in agreement with Harvey et al.'s (2001) conclusions from seismic receiver function analysis. In the southern domain, the genesis of the box folds requires steeply south dipping controlling faults at a location that is consistent with the observed normal faults. Furthermore, the offshore observations demonstrate that as the Mesozoic syn-rift displacement is at least 12,000 m, the minimum depth to the base of the Peninsula Formation is at least 14,000 m. Therefore, if the Cape Fold Belt is underlain by a mega-detachment, the surface onto which the faults assimilate has a depth of at least 14,000 m. In addition, given that the observed Mesozoic syn-rift fault throws increase towards the south (Port Elizabeth, Gamtoos and Plettenberg Faults respectively), this may be a consequence of the south dipping nature of a mega-detachment (Fig. 8).

It has been demonstrated that the Cape Fold Belt underwent compression on high angle controlling faults in the south and that these faults were then reactivated in extension. Given that the Cape Supergroup was deposited onto a passive margin it is likely that the margin was controlled, at least during its rift phase, by active normal faults. We speculate that given the high angle faults that generated the compressisonal geometries are likely to have inherited their dips from pre-existing structures (Williams and Powell, 1989), these reverse faults may represent normal faults that were active during the deposition of the Cape Supergroup margin (Fig. 8). Although the Cape Supergroup is not a uniform thickness across the fold belt, further studies would have to be undertaken to investigate whether the thickening is associated with the presence of the fault structures as predicted by this model.

From the available data it remains enigmatic as to what controls the variation in structural domains. The transition between structural domains is, however, co-incident with the presence of the Southern Cape Conductive Belt and the Beattie Anomaly (De Beer, 1983; Pitts et al., 1992) which has led previous studies to infer that it is associated with variations in lithospheric strength or thickness between the north and south (Söhnge and Hälbich, 1983; De Wit and Ransome, 1992; Hälbich, 1993; Harvey et al., 2001).

6. Discussion

6.1. Applicability of thin- and thick-skinned tectonic models

Compressional orogenies are commonly attributed to either thin- or thick-skinned tectonic models (Fig. 1) with the former involving deformation of cover sequences on ramp and flat thrust planes without basement involvement, while the latter involves crustal scale ramps and basement interaction (e.g. Coward, 1983). Here we discuss the applicability of these two models to the Cape Fold Belt.

As has been discussed, the northern domain is characterised by many features commonly associated with thin-skinned tectonics, and previous workers have made this comparison (Booth and Shone, 1999; Booth, 2001). However, there are other features typical of thin-skinned models that are not evident, including significant klippen, imbrications, varying degrees of metamorphism, juxtaposition of out-of-sequence units, requirement for significant duplication of cover sequences at depth (e.g. Boyer and Elliott, 1982; Butler, 1982; Coward, 1984; Butler et al., 1986; Dewey et al., 1986; Le Fort, 1986; Vann et al., 1986; Hossack, 1983; Mitra, 1990; Spring and Crespo-Blanc, 1992; Mercier et al., 1997; Philippe et al., 1998). The absence of such features may be a consequence of the relatively small amount of shortening that the Cape Fold Belt has undergone (~20%). However, the cross-sections predict that it is more likely to be associated with the nature of the detachment, which is considered to be within the Pre-Cape unit below the Cape Supergroup, and hence within the basement of the fold belt. Therefore, superficially, the northern domain resembles thin-skinned tectonics, including a shallow detachment structure, however, the cross-sections imply involvement of basement at depth. In comparison, the southern domain, with steeply dipping controlling faults and Pre-Cape-, or basement-cored anticlines conforms more with a thick-skinned model. However, this is also problematic because Coward's (1983) definition of thick-skinned tectonics requires near vertical movements of crust associated with continuation of such steep structures to depth. Such a model is, therefore, incompatible with the southern structural domain if the high-angle controlling faults are considered to coalesce onto the mega-detachment.

Associated with this is the definition of basement. Within the fold belt the oldest unit to crop out is the Proterozoic Pre-Cape, therefore, it has been considered as the basement. However, given that previous workers have documented that the Pre-Cape units are non-crystalline and were deposited into extensional basins that are intimately linked to the regional heterogeneity, and may well be controlled by extensional faults that reactivate a possible mega-detachment, the Pre-Cape units could therefore be considered to be non-basement units. An important distinction may, therefore, exist between crystalline basement, which in South Africa, would be considered to be the 2.1–1.0 Ga Namaqua-Natal belt and the overlying sequences including the 1.0–0.5 Ga Pre Cape unit as cover.

Therefore, in settings such as the Cape Fold Belt the application of thin- or thick-skinned tectonic models may not be directly applicable. A more useful way of describing the deformation is to classify it as shallow dipping controlling structures in the north and progressively steeper dipping towards the south that may décolle onto a regionally south dipping mega-detachment (Hälbich, 1993; De Wit and Ransome, 1992; Söhnge and Hälbich, 1983).

6.2. Fault dips that are susceptible to inversion

The Mohr-Coulomb criterion of failure within homogenous crust predicts that faults with typical values of internal friction will develop at c. 30° to σ1, hence, reverse faults have an idealised dip of approximately 30° and normal faults of 60°. If the rock volume is heterogeneous, as is common within the continental lithosphere, then a pre-existing structure will be reactivated if the applied stress is less than the critical stress required for the formation of a new fracture (Ranalli, 2000). However, a variety of parameters, including fault dip, stress orientation with respect to pre-existing fabric, pore pressure and fluid involvement, frictional coefficient of fault plane, thermal properties, are considered to influence whether a pre-existing structure will be reactivated or not (e.g. Ivins et al., 1990; Sibson, 1985, 1995; Forsyth, 1992; Huyghe and Mugnier, 1992; Van Wees and Beekman, 2000; Ranalli, 2000). The large number of variables has led to a proliferation of observed and theoretical concepts at which faults will be inverted rather than cross cut, such as the suggestion by Faccenna et al. (1995) that pre-existing faults will be ignored if fault dip is less than 32°, partly reactivated if between 32° and 41° or totally reactivated for greater than 41°. In addition, when steeply dipping faults are inverted in a reverse sense, new low dip footwall cut off thrust are expected to develop, for example footwall cut-offs associated with inverted faults dipping 50–60° in sand box modelling (McClay, 1989; Buchanan and McClay, 1991).

This study demonstrates that a suite of faults with dips from 24–60° can be inverted in both an extensional and compressional sense. Even with the steepest faults, there is no evidence of footwall cut-off faults. Further studies would be required to determine why such faults are able to be inverted, including investigating pore pressure, fault friction, or whether the basement lihology is rheologically strong enough. We, therefore, suggest that given the correct conditions, faults of very variable dips can be reactivated.

6.3. Multiple inversion events

Given the complexity of deformation that is inherent in regions that have undergone inversion, it is commonly difficult to differentiate between various stages of deformation. A number of settings have demonstrated the difficulty, and often controversy, in determining the relative ages of faults in complex fault systems; such difficulties are compounded when the relative timing of fault reactivation is considered (e.g. Williams and Fischer, 1984; Butler, 1989; Coryell and Spang, 1988; O'Dea and Lister, 1995; Constenius, 1996; Travelli, 1999; Scisciani et al., 2002; Calabrò et al., 2003; Butler et al., 2004).

Southern Africa, therefore, provides a rare example in which the same structures are observed to have been reactivated during a number of phases and that the timing can be demonstrated. As discussed, it is speculated that the passive margin faults during the Cape Supergroup deposition were reactivated as high angle reverse faults during the Permian Cape Orogeny prior to further reactivation as extensional faults during the Mesozoic. In addition to these episodes, it has been documented that the Pre-Cape units were deposited into extensional basins that have very similar geometries to the subsequent extensional systems (e.g. Tankard et al., 1982; Dingle et al., 1983; Gresse, 1983). It is, therefore, possible that the faults associated with the Pre-Cape unit were reactivated during the Cape Supergroup passive margin, although such an assertion requires a much better understanding of the limited Pre-Cape unit.

An important observation from the Mesozoic reactivation of the Permian compressional faults is that only selective reactivation has occurred. There are a number of compressional faults inferred from the presence of box folds (Fig. 8) that have not undergone extensional reactivation during the Mesozoic. From the present data it is unclear why this has occurred, however, it may be a consequence of the orientation of the extensional stress during the development of Mesozoic basins (Paton and Underhill, 2004).

7. Conclusions

Through integrating surface geology and structural reconstructions with subsurface data that constrain the dimensions and geometry of the underlying fault systems, cross-sections through the Cape Fold Belt of southern South Africa have been constructed. These sections demonstrate the similarity between the controlling structures required from the observed compressional and extensional deformation while also demonstrating the significant variation in the deformation across the fold belt.

Our model for the development of the fold belt proposes that faults active in an extensional sense during the Cambrian-Devonian Cape Supergroup passive margin episode were reactivated as reverse faults during the Permian compression. In the south these faults are high angle structures that during compression resulted in the generation of box folds and deformation akin to a thick-skinned style of tectonics. In contrast, faults in the north are shallower dipping, resulting in a compressional geometry predicted by thin-skinned tectonic models. Subsequent negative inversion during the Mesozoic, resulted in extensional geometry that reflects this variation, with high angle normal faults in the south and shallower dipping normal faults in the north.

We conclude:

1) The southern Cape region of South Africa has undergone at least two episodes of structural inversion: first in compression, forming the Cape Fold Belt; and subsequently extension in the Mesozoic. During both episodes fault geometry has remained consistent.
2) Structures with dips that vary between 24° to 60° have been reactivated in both extension and in compression, which is in contrast to simple fault reactivation models.
3) The Cape Fold Belt cannot be classified as either a thin-skinned or thick-skinned tectonic end-member as these characterisations do not adequately described the features that have been observed. Instead, it is more useful to describe the deformation as being controlled by a south dipping mega-décollement that exhibits aspects of both end-members.

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